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Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada

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Abstract:

The water balance and runoff regime of a 55 ha aspen-forested headwater catchment located on the Boreal Plain, Alberta, Canada (55·1 °N, 113·8 °W) were determined for 5 years following a partial timber harvest. Variability in precipitation provided the opportunity to contrast catchment water balances in relatively dry (<350 mm year^{-1}), wet (>500 mm year^{-1}), and average precipitation years. In most years, the catchment water balance was dominated by soil water storage, evapotranspiration losses, and vertical recharge. In 1997, despite near-average annual precipitation (486 mm), there was significant runoff (250 mm year^{-1}) with a runoff coefficient of 52%. A wet summer and autumn in the preceding year (1996) and large snow accumulation in the spring (1997) reduced the soil water storage potential, and large runoff occurred in response to a substantial July rainfall event. Maps of the surface saturated areas indicated that runoff was generated from the uplands, ephemeral draws, and valley-bottom wetlands. Following 1997, evapotranspiration exceeded precipitation and large soil water storage potentials developed, resulting in a reduction in surface runoff to 11 mm in 1998, and <2 mm in 1999–2001. During this time, the uplands were hydrologically disconnected from ephemeral draws and valley-bottom wetlands. Interannual variability was influenced by the degree of saturation and connectivity of ephemeral draws and valley wetlands. Variability in runoff from tributaries within the catchment was influenced by the soil water storage capacity as defined by the depth to the confining layer. An analysis of the regional water balance over the past 30 years indicated that the potential to exceed upland soil water storage capacity, to connect uplands to low-lying areas, and to generate significant runoff may only occur about once every 20 years. The spatial and temporal variability of soil water storage capacity in relation to evaporation and precipitation deficits complicates interpretation of forest harvesting studies, and low runoff responses may mask the impacts of harvesting of aspen headwater areas on surface runoff in subhumid climates of the Boreal Plain.

KEY WORDS aspen forest; Boreal Plain; evapotranspiration; forest harvest; runoff; soil storage; surface saturation; water balance

INTRODUCTION

The glaciated Boreal Plain of western Canada is experiencing accelerated rates of timber harvesting activities, oil and gas exploration and extraction, mining, and recreation (AEP, 1998). Further, an increase in the demand for aspen has occurred owing to the recent economic importance of aspen for pulp and paper. Large portions of the Boreal Plain are covered with extensive stands of aspen (Hall *et al*., 1997). Increased understanding of the hydrology of aspen-dominated regions is required to predict the degree to which environmental changes on this region of the Boreal Forest will affect surface waters (Buttle *et al*., 2000).

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To date there has been little process-based research addressing the potential impacts of timber harvesting on runoff from forests in the subhumid region of the Boreal Plain (Buttle *et al*., 2000). Recently, studies conducted in humid regions of the Boreal Shield have accumulated evidence that timber harvesting results in increased base and peak flows resulting from decreased snow and rainfall interception, and increased infiltration (Buttle *et al*., 2000). There, the soil water storage capacity is regularly exceeded, and the geology facilitates lateral runoff that increases the potential impacts of forest removal (McDonnell, 2003). The drier climate and deep glaciated substrate of the Boreal Plain most likely result in complex surface–groundwater interactions, and the dominance of soil water storage and evapotranspiration (Devito *et al*, 2000; Woo *et al*., 2000; Grayson and Western, 2001). A large portion of annual precipitation falls during the summer in continental western Canada, and the synchronization of rainfall with peaks in evapotranspiration further reduces the potential for runoff. This contrasts the humid regions of Canada, where the dominance of snowmelt regimes exacerbate the impacts of harvesting (Buttle *et al*., 2000). Furthermore, natural variability in climatic conditions has been shown to confound the influence of timber harvesting on surface water quantity and quality in both humid and subhumid regions of the Boreal Forest (Buttle and Metcalfe, 2000; Devito *et al*., 2000).

Upland and low-lying valley or wetland riparian areas may behave independently in runoff generation (Grayson and Western, 2001; McDonnell, 2003), and their behaviour is not well documented in the Boreal Plain. Impermeable layers of frozen soils in valley-bottom wetlands have been shown to reduce the soil water storage capacity and to serve as preferential pathways for meltwater runoff in many subhumid or semi-arid, subarctic and arctic catchments (Glenn and Woo, 1997; Carey and Woo, 1998). These studies clearly show that large differences in rainfall or meltwater absorption can be expected between well-drained uplands and poorly drained lowlands and/or ephemeral draws with extensive frozen soils. Thus, only portions of the catchments may act as source areas for runoff generation (Woo *et al*., 2000; McDonnell, 2003).

Predicting the potential impacts of forest harvesting on runoff generation in aspen forests on the Boreal Plain requires a sound understanding of the natural variation in aspen forest water balance and runoff regimes. For example, in both the Wolf Creek watershed in the Yukon on the Boreal Cordillera ecozone (Carey and Woo, 1999), and in the aspen forest on prairie sloughs in the Prairies Ecozone (van der Kamp *et al*., 2003), water is either intercepted or infiltrated and stored until subsequent evapotranspiration. As such, significant runoff generation during summer storms is restricted. Given that these ecozones may represent end members for climatic conditions on the Boreal Plain, it is predicted that there will be no or low runoff from aspen forests on the Boreal Plain (Blanken *et al*., 2001). Thus, the impacts on runoff generation may be weakened if there is a significant soil water storage capacity compared with the difference in precipitation and evapotranspiration. However, the impacts on runoff generation may be intensified if there is significant formation of frozen soils that reduce the effective soil water storage capacity to the point where it cannot accommodate the increase in water resulting from snowmelt or spring rainfall (cf. Glenn and Woo, 1997; Carey and Woo, 1999; Buttle *et al*., 2000). Long-term studies in aspen-dominated catchments are imperative to identify the potential impacts of timber harvesting and other disturbances on hydrologic processes in forests on the Boreal Plain (Buttle *et al*., 2000).

Herein, we report on 5 years of data quantifying the natural variability in runoff generation from a headwater catchment representative of aspen-dominated, glaciated, subhumid regions of the Boreal Plain ecoregion. This study is part of an integrated research programme examining landscape position as an organizing principle to predict the potential sensitivity of lakes and wetlands to disturbances in order to provide a framework for forest management strategies (Buttle *et al*., 2000; Devito *et al*., 2000). A test of the landscape position hypothesis requires a firm understanding of the location and extent of runoff-generating source areas that export water from contributing hillslopes to aquatic ecosystems. Concurrent measures of lithology, ground water-table dynamics, soil water content, and runoff within the catchment were conducted to determine: (1) source areas contributing to runoff; (2) runoff generation mechanisms; and (3) regulation of runoff generation by precipitation, soil water storage capacity, soil water content, and surface saturated areas (SSAs) in the catchment. This knowledge, in conjunction with a comparison of harvested and non-harvest portions of the catchment, were then used to infer the influence of timber harvesting on runoff on the Boreal Plain.

STUDY SITE

This study was conducted in a zero-order catchment draining into Moose Lake (55°97′58″N, 111°46′17″W) located approximately 250 km northeast of Edmonton, Alberta, Canada (Figure 1). The headwater catchment is located in the Mixed-Wood Boreal ecoregion (Rowe, 1972) of the Boreal Plain ecozone (Ecological Stratification Working Group, 1996). The climate is continental, characterized by cold winters and warm summers. The average (1970 to 2000) annual air temperature is 1·5 °C, with monthly ranges from −16·7 °C (January) to 16·3 °C (July). The average annual potential evapotranspiration (538 mm) exceeds annual precipitation (462 mm), with the majority of the rainfall (>60%) coming in June, July, and August, largely as convective storms of high intensity and short duration (Devito and Fraser, 2004). The region is often covered in snow by early November. Precipitation during the winter months represents <25%(<100 mm) of the average annual precipitation, and sublimation results in only small accumulations of snow (Pomeroy *et al*., 1997). The average runoff from the nearby Logan River watershed (425 km^2) is about 128 mm $year^{-1}$ (Devito and Fraser, 2004).

The catchment (52·4 ha) is drained by an ephemeral draw (no channel incision occurs) originating from a minerotrophic, mixed-wood wetland (2·1 ha) (Figure 2). Three ephemeral streams (subcatchments A (14·6 ha), B (5·5 ha), and C (18·3 ha)) drain into the wetland. The catchment has approximately 50 m of relief, with an average slope of 8%. Clay-rich glacial till is about 80 m thick (Tokarsky and Epp, 1987; Devito *et al*., 2000). There is a surface ablation till that ranges from 0·5 m to over 4·5 m in the valley-bottom wetlands and ephemeral draws, and 0·5 to 2 m in the uplands. In the surface ablation till, saturated hydraulic conductivities range from 10^{-8} to 10^{-5} m s^{-1}, with slightly higher hydraulic conductivities in the ephemeral draws reflecting a higher frequency of sand layers (Devito and Fraser, 2004). Gray Luvisols and Eutric Brunisols have developed from the surface ablation till in the uplands. Gleysols and organic soils (0·5 to 2·5 m), occasionally mixed with sand and silt, occur in the valley-bottom wetland, with saturated hydraulic conductivities ranging from 10^{-7} to 10^{-6} m s^{-1} (Kalef, 2002; Devito and Fraser, 2004). The saturated hydraulic conductivities of the compacted confining layer below the ablation till range from 10^{-9} to 10^{-8} m s^{-1}.

The forest is dominated by trembling aspen (*Populus tremuloides* Michx.) with some paper birch (*Betula papyrifera* Marsh.) and balsam poplar (*Populus balsamifera* L.), and occasional small stands of white spruce

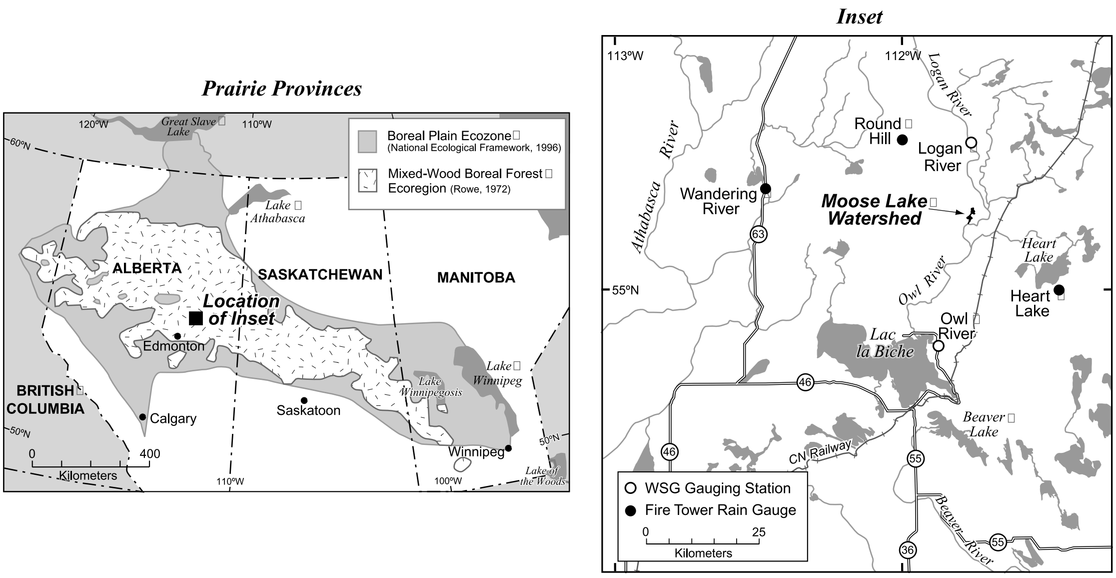

Figure 1. (a) Location of Moose Lake in relation to the Mixed-Wood Boreal Forest ecoregion and the Boreal Plain ecozone of Canada

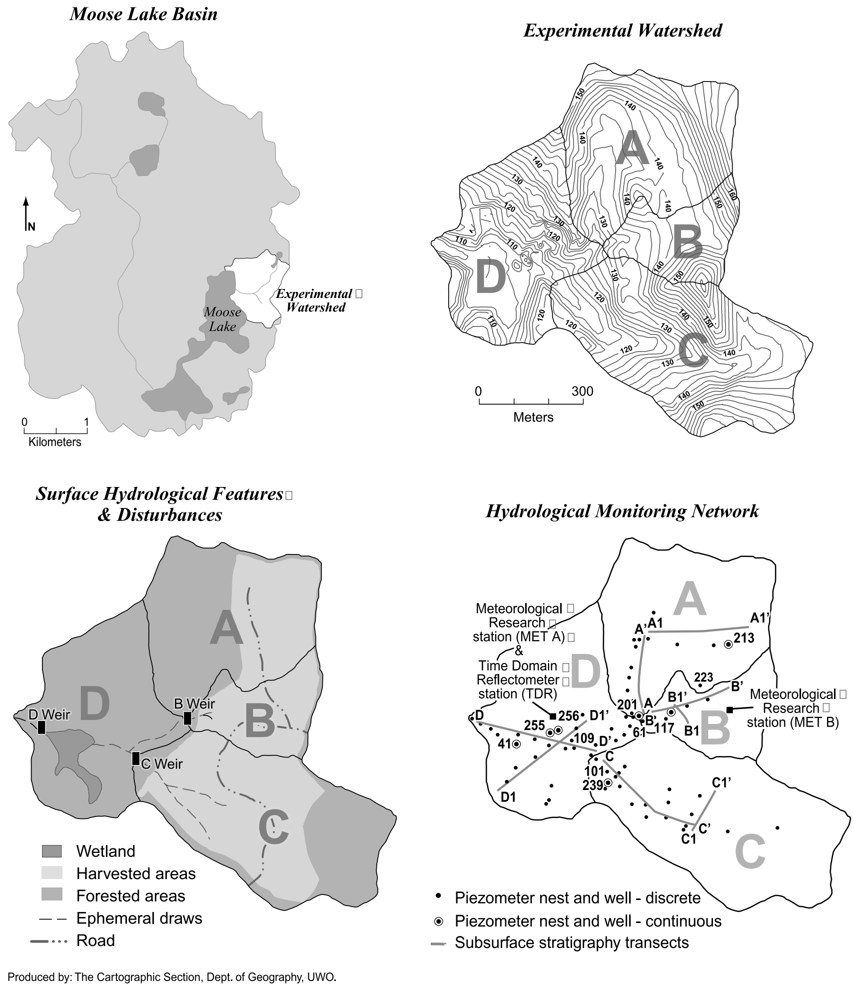

Figure 2. Location of the study catchment draining into Moose Lake showing the topography, the locations of subcatchments A, B, C, catchment D and weirs, the valley-bottom wetlands and ephemeral draws, the harvested areas, and the surface and subsurface instrumentation

(*Picea glauca* (Moench) Voss). The wettest areas are characterized by paper birch, balsam poplar, and speckled alder (*Alnus rugossa* (Du Rois) Spreng.).

Comparison of harvested and forested subcatchments

In January 1997, approximately 26·6 ha (51%) of the catchment was logged for trembling aspen, representing 34% of subcatchment A, 88% of subcatchment B, and 73% of subcatchment C (Figure 2). No logging occurred in the lower reaches of the catchment. At the time of harvest, conducted with a faller-buncher, the frozen ground reduced the potential impacts to soil structure. Soil samples from 1998 to 2001 indicate little difference in forest floor depth and soil bulk density in harvested and forested areas of the catchment, with the exception of roadways (Devito, unpublished data). Rapid forest regeneration occurred in the cut-blocks, with suckers of trembling aspen exceeding 0·3 m in height at the end of the first growing season (1997) and 1 m in height by the end of the second growing season (1998). Slow forest regeneration occurred on the roads, with minimal forest growth evident 5 years following the timber harvest.

Stream flow monitoring was initiated following a harvest; therefore, no pre-cut data (before winter 1996–97) are available. However, the forest harvest permitted a comparison of predominantly forested and harvested tributaries of the main ephemeral draw to assess the influence of tree harvesting on runoff yields. The subcatchments were gauged in May 1998. Thus, the potential influence of harvesting following one full growing season of regeneration of aspen headwaters can be assessed.

METHODS

The water balance per unit area (millimetres per year) of catchment or subcatchment was determined for the water year, 1 November to 31 October, and is expressed as

$$r = P - \mathrm{AET} - R_{\mathrm{SF+SSF}} - \mathrm{GrW_r} - \Delta S_{\mathrm{USZ+SZ}} \tag{1}$$

where r is the residual error, P is precipitation, AET is actual evapotranspiration, $R_{\mathrm{SF+SSF}}$ *is* surface flow (SF) and subsurface flow (SSF) runoff, $\mathrm{GrW_r}$ is groundwater recharge, and $\Delta S_{\mathrm{USZ+SZ}}$ is the change in the soil water content in the unsaturated zone (USZ) and the saturated zone (SZ). The following sections describe how each component of the water balance was estimated.

Precipitation

Meteorological recording stations were installed in the forested (MetA) and harvested (MetB) areas in the catchment that recorded P(mm day^{-1}), air temperature, and soil temperatures at 10 and 50 cm depth (Figure 2). Missing data were interpolated from data collected at four nearby (<25 km) meteorological recording stations. If the air temperature was $<2\,^{\circ}$C, then the precipitation was assumed to be snow. Snow depth, snow water equivalent, and snow density were determined in late February or early March of each year at five sampling stations along 16 transects evenly distributed within the forested and harvested areas of the catchment.

Potential and actual evapotranspiration

Potential evapotranspiration (PET) (mm day^{-1}) was estimated using the Thornthwaite (1948) method. A 6 m tower with a net radiometer (Q^*), soil heat flow transducer at 8 cm depth ($G_{8\mathrm{cm}}$), soil temperature probes at 2 and 6 cm depths and a soil moisture probe at 2·5 cm depth were installed in early spring 2001 at MetB (Figure 2). Additionally, wind speed and direction measurements at 4 m height and air temperature and relative humidity (RH) measurements at 2 m were taken at this site. For 2001, daily PET was estimated using Priestley–Taylor and Penman–Monteith methods (Oke, 1987). For the Priestley–Taylor method, the coefficient α (which is equal to the ratio of PET to equilibrium evapotranspiration) was assumed to be 1·26, which is appropriate for regions with no or low advective conditions (Oke, 1987). For the Penman–Monteith method, the bulk density of the sandy soil was assumed to be 1800 kg m^{-3} and the heat capacity of dry mineral soil was assumed to be 850 J kg^{-1} K^{-1}. From June to September 2001, monthly estimates of PET based on the Thornthwaite method were within 10% of monthly estimates of PET based on the Priestley–Taylor and Penman–Monteith approaches, justifying the use of the Thornthwaite formula for earlier years.

Direct estimates of AET (mm day^{-1}) were not made. An approximation of AET was based on the water balance:

$$\mathrm{AET_{WaterBalance}} = P - R_{\mathrm{SF+SSF}} - \mathrm{GrW_r} - \Delta S_{\mathrm{USZ+SZ}} \tag{2}$$

A second approximation of AET was based on the seasonal P (1 May to 15 October) and the integration of seasonal changes in the soil water content in the top 1 m of the soil at the meteorological recording site (MetA) (Equation (3)):

$$\mathrm{AET_{Meteorology}} = P - \Delta S_{\mathrm{USZ+SZ}} \tag{3}$$

This net rate of AET includes summer interception and assumes both minimal changes in soil moisture below 1000 mm, where root growth is reduced, and a similar distribution of soil characteristics throughout the catchment. $AET_{Meteorology}$ was determined independently of the water balance and, thus, it was used in the calculation of the water balance for the catchment.

Surface and subsurface flow

Prior to the installation of weirs, R_{SF}(mm day^{-1}) was determined by manual measurements of stage height and spot runoff estimates of water flowing in naturally confined areas at the outflow of the catchment and at subcatchments A, B and C. In 1998, following the installation of weirs, R_{SF} was continuously monitored using stage height versus runoff rate relationships established for 60° or 90° V-notch weirs at the outflow of the catchment and subcatchments B and C (Figure 2). Weirs designed to capture R_{SF} were inserted to a depth of 30 to 50 cm in the substrate and sealed with plastic, sheet metal, and clay substrate or bentonite clay. Well and piezometer nests were installed above and below each weir and used to estimate R_{SSF}(mm day^{-1}). Soil transmissivity was determined from the saturated hydraulic conductivities of each soil texture and the distribution of each soil texture beneath each weir. It was assumed that R_{SSF} flow was confined to the soil volume between both breaks in slope in the valley containing the weirs in each subcatchment.

Groundwater recharge

A network of wells and piezometers was installed throughout the catchment (Figure 2). Wells were fully perforated and wrapped in 'well socks' prior to installation. Piezometers were constructed from 2 cm ID PVC pipe and coupled to 0·2 m slotted and shielded PVC heads. Wells were inserted into pre-bored holes extending into the underlying confining material and back-filled with aggregate. Piezometers were inserted into pre-bored holes at 1 to 1·5 m depth intervals until the confining material was intercepted and back-filled with aggregate (shielded head) and bentonite (piezometer shaft). GrW_r(mm day^{-1}) was estimated using the Darcy approximation (Fetter, 1999), where subsurface stratigraphy was estimated from soil cores collected at each piezometer nest and hydraulic conductivities were estimated from bail tests of the piezometers (Hvorslev, 1951). The rate of groundwater recharge was determined from the vertical hydraulic gradients between the base of the surface ablation till and in the deeper confining layer at piezometer nests in the uplands and lowlands of each subcatchment.

Change in soil water content

Changes in S_{USZ}(mm day^{-1}) were determined at MetA using time-domain reflectometry (TDR) probes inserted at the Organic–Ae interface (8 cm), at the sand–clay interface (27 cm), in the clayey-sand lens (35 cm), and in the sand below the clayey-sand lens (60 cm) on the hillslope adjacent to the valley-bottom wetland. (Figure 2). Changes in S_{SZ} were determined from water table fluctuations that were monitored continuously in seven wells: three in the valley-bottom wetland and four in the uplands (Figure 2). Specific yields were estimated from the relationships between precipitation and water table elevation, and/or literature values for soil texture (Saxton *et al.*, 1986).

SSAs in the catchment were mapped several times each year relative to a geographically referenced grid network. An area was considered saturated if moderate pressure applied by a boot forced gravimetric water from the soil or if standing water was observed. Maps of the areas of saturation were generated, digitized, and then used to estimate the total SSA both within the subcatchment and within the entire catchment.

RESULTS AND DISCUSSION

Annual water balance

Natural variability in precipitation during the 5 years of the study provided the opportunity to contrast water balances in years with relatively wet, dry, and average annual precipitation (Table I). Among the

Table I. Annual water balance for the study catchment, the regional runoff of Logan River, and the runoff from subcatchments A, B and C, based on the hydrologic year from 1 November to 31 October. Complete records for hydrologic year 1997 and 1998 were not available for subcatchments A, B, and C. R for Logan River, 2001, not available

Year	P (mm)	PET (mm)	AET[a] (mm) WB	AET[a] (mm) TDR	R_{SF} (mm)	R_{SF+SSF} (mm)	RC (%)	GrW recharge[b] (mm)	S_{GrW}[c] (mm)	S[d] (mm)	S_c[e] (mm)	Residual error[f] [mm (%)]
Study catchment D												
1997	486	484	226	—	250	251	52	9	—	—		—
1998	329	566	364	—	11	12	4	10	−56 to −58	—	−56 to −58	—
1999	318	526	319	249	2	2	1	9	11 to −29	−3	8 to −32	70 [22]
2000	529	495	439	533	2	2	0	10	55 to 57	22	77 to 79	−94 [−18]
2001	444	522	487	453	0	1	0	8	−11 to −29	−31	−42 to −60	34 [8%]
Logan River (regional)												
1997	486	—	192	—	—	294	60	—	—	—	—	—
1998	329	—	255	—	—	74	22	—	—	—	—	—
1999	318	—	257	—	—	61	19	—	—	—	—	—
2000	529	—	409	—	—	120	23	—	—	—	—	—
2001	444	—	444	—	—	—	—	—	—	—	—	—
Subcatchment A												
1999	318	—	320	—	0	2	1	8	—	—	—	—
2000	529	—	418	—	0	2	0	31	—	—	—	—
2001	444	—	474	—	0	1	0	20	—	—	—	—
Subcatchment B												
1999	318	—	273	—	10	54	17	3	—	—	—	—
2000	529	—	369	—	36	80	15	2	—	—	—	—
2001	444	—	437	—	13	57	13	1	—	—	—	—
Subcatchment C												
1999	318	—	327	—	1	2	1	1	—	—	—	—
2000	529	—	448	—	2	2	0	1	—	—	—	—
2001	444	—	492	—	2	2	1	1	—	—	—	—

[a] WB: derived by water balance; TDR: derived from summer soil moisture depletions relative to precipitation.
[b] Recharge to deep groundwater.
[c] Groundwater storage.
[d] Storage in soil.
[e] Total change in catchment soil storage.
[f] That portion of the budget that is not accounted for based on AET determined by TDR.

5 years, P varied by 211 mm year^{-1} and PET varied by only 82 mm year^{-1}. PET equaled P in 1997, but exceeded P in 1998, 1999, and 2001, and P was greater than PET in 2000. We recognize the inherent errors in estimating AET; however, both independent measures using water balance calculations and soil moisture changes indicated that AET $<$ PET in most years, and that AET represented the largest export of water from the catchment. With the exception of 1997, changes in AET were directly related to changes in P, indicating that AET was limited by the availability of water. AET was near its potential rates in 2000 and 2001. Runoff varied by over 250 mm year^{-1}, but this variability was not related to P. The greatest R(251 mm year^{-1}) and runoff coefficient (RC, 52%) occurred in 1997, a year with average P(486 mm year^{-1}). After 1997, R decreased from 12 mm year^{-1} in 1998 (RC = 4%), to 1 mm year^{-1} in 2001 (RC <1%). Despite the relatively large P in 2000 (529 mm year^{-1}), the catchment was unable to recharge and the runoff remained low. In the last 3 years of the study, groundwater seepage rates were higher than R, ranging from 8 to 10 mm year^{-1}, and represented a small but consistent export of water from the catchment. Soil and groundwater storage capacities were also higher compared with changes in R, indicating that the soil water storage capacity was rarely exceeded.

To gain insight into the hydrologic dynamics of 1997, we conducted an analysis of longer trends in the water balance that could be computed for the nearby Logan River watershed. Extrapolation of the water balance of the catchment based on the Logan River watershed was considered reasonable given the similarity in the trends, if not magnitude, of the water balance components during the study period (Table I). The longer trend in the water balance of the Logan River watershed suggests that the runoff response in 1997 was influenced by soil water recharge during 1996. The excess of *P* over PET was larger in 1996 (120 mm year^{-1}) than in 1997 (2 mm year^{-1}). However, *R*(211 mm year^{-1}) and RC (37%) were smaller in 1996 than the *R*(251 mm year^{-1}) and RC (52%) in 1997. Prior to 1996, PET exceeded *P* for about 10 consecutive years (Figure 3). During this time, the region was unable to recharge, and *R* and RC of the Logan River were small and similar to the *R* and RC after 1997. This analysis suggests that the *R* from the catchment in 1997 was exceptional, the largest in about 20 years, and that the *R* and RC from the catchment during the period from 1998 to 2001 was common (Figure 3). The low RCs for the catchment are similar to RCs reported in warmer but subhumid climates, where RCs of 7% to 25% are common (Burch *et al*., 1987; Pinol *et al*., 1991). The high RC in 1997 is similar to more humid climates (Likens *et al*., 1977; Allen and Roulet, 1994), but was short lived.

As expected in subhumid regions, there was a poor relationship between *P* and *R* for both the 5 years of catchment measurements and the 20 years for the Logan River watershed ($r^2 = 0{\cdot}11$, $p > 0{\cdot}10$). This situation contrasts with the poor relationship between *P* and PET, but the good relationship between *P* and *R*, in humid regions (Everson, 2001).

Seasonal water fluxes

Differences in the timing and magnitude of seasonal *P* relative to PET within a year are important in explaining the large variation in runoff responses among the years. In most years, the precipitation depths from November to April were small (<100 mm; Figure 4), resulting in little or no autumn or spring runoff from the catchment (Figure 5, Table I). This contrasts with conditions in humid regions, where autumn rains typically recharge catchments and spring snowmelt and rains produce from 50 to 70% of the annual *R* (e.g. Devito *et al*., 1996).

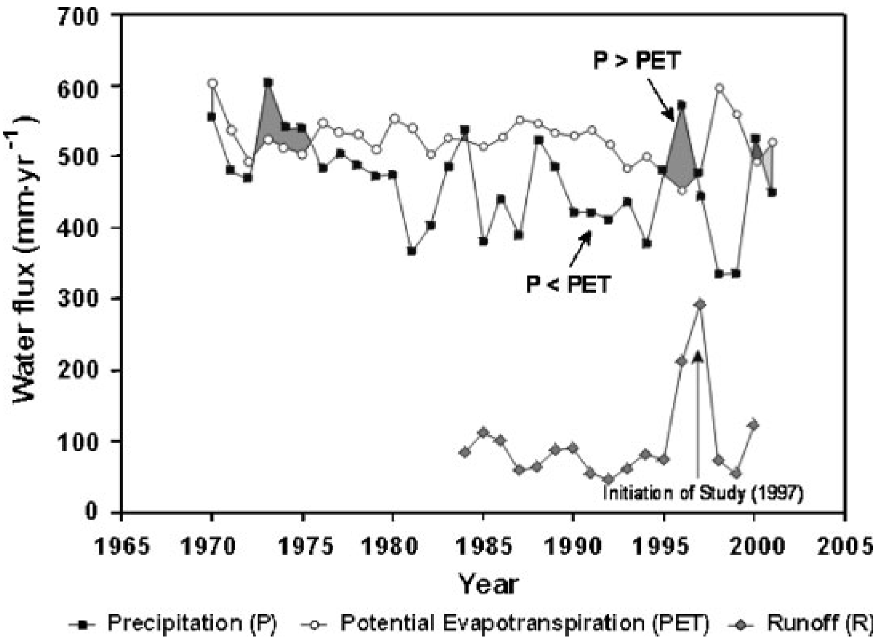

Figure 3. Thirty-year (1970–2000) summary of precipitation *P* and PET for the study area, and Logan River runoff *R* from 1984 to 2000

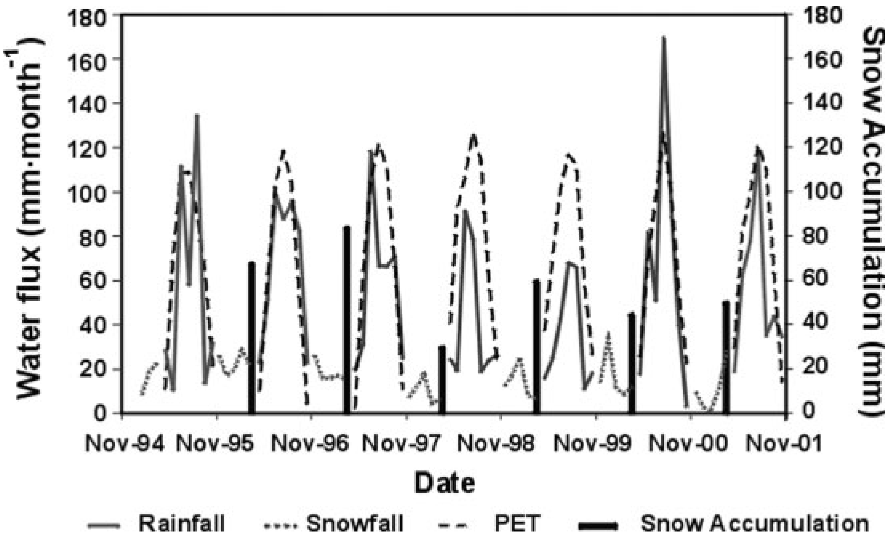

Figure 4. Monthly (1996–2001) precipitation (P, snow versus rain), accumulated precipitation as snow by 15 March, and PET for the Moose Lake study catchment

In the catchment, the majority of runoff occurred in the summer. No runoff data were available for the catchment prior to 1997, and only instantaneous runoff data were available for 1997. Based on data from the Logan River watershed, the relatively large runoff that occurred in 1997 was associated with a relatively large snow accumulation followed by a summer where $P \cong$ PET (Figure 4) with frequent storms $\geq$20 mm day^{-1} (Figure 5). In 1996, late summer–early autumn rainfall events resulted in conditions where $P >$ PET and major R occurred (Figure 4). Following the relatively wet autumn, large snow accumulation probably reduced the soil water storage capacity and/or increased soil freezing, resulting in increased runoff from snowmelt in spring of 1997. The wet spring was followed by a wet summer (i.e. several storms $\geq$20 mm day^{-1}) during which monthly $P >$ PET. For example, a single rainfall event from late on 18 June to early on 20 June produced 70 mm of precipitation in the catchment in less than 48 h. There are considerable errors in estimating daily runoff from the instantaneous measurements that were available in 1997; nevertheless, the largest daily instantaneous measures and RC (>35%) for both the study catchment and Logan River were associated with this rainfall event. This indicates that the soil water storage capacity in the clay-rich soils was exceeded, and large areas of the catchment contributed to the storm runoff.

In contrast to 1997, most of the summer months in the other years had $P <$ PET, $R < 1$ mm day^{-1}, and RC < 1% (Figure 5, Table I). Plots of rainfall versus peak runoff illustrate the non-linear rainfall-runoff response and dependence on antecedent soil moisture conditions (Figure 6). For example, in 1998, a 60 mm rainfall following a 40 mm event 2 weeks earlier generated only a moderate runoff response (2·6 mm day^{-1}), the peak runoff for the year. However, soil water content appears to have been rapidly depleted as, about 1 week after the peak runoff, a 30 mm rainfall event generated only a 0·8 mm day^{-1} runoff response. In 1999, an isolated rainfall event of almost 30 mm did not generate surface runoff. In 2000, three successive rainfall events $\geq$30 mm occurred within about 10 days, but generated only a relatively minor runoff response (<1 mm day^{-1}). Generally, little or no runoff occurred during isolated rain events ($\leq$30 mm), and peak runoff occurred only when successive storms ($\geq$30 mm day^{-1}) occurred within about 10 days. The summer (and thus annual) rainfall and runoff patterns on the subhumid Boreal Plain are similar to summer rainfall and runoff on the humid Boreal Shield (Allan and Roulet, 1994), where many catchments have little or no runoff during this season (e.g. Devito *et al*., 1996).

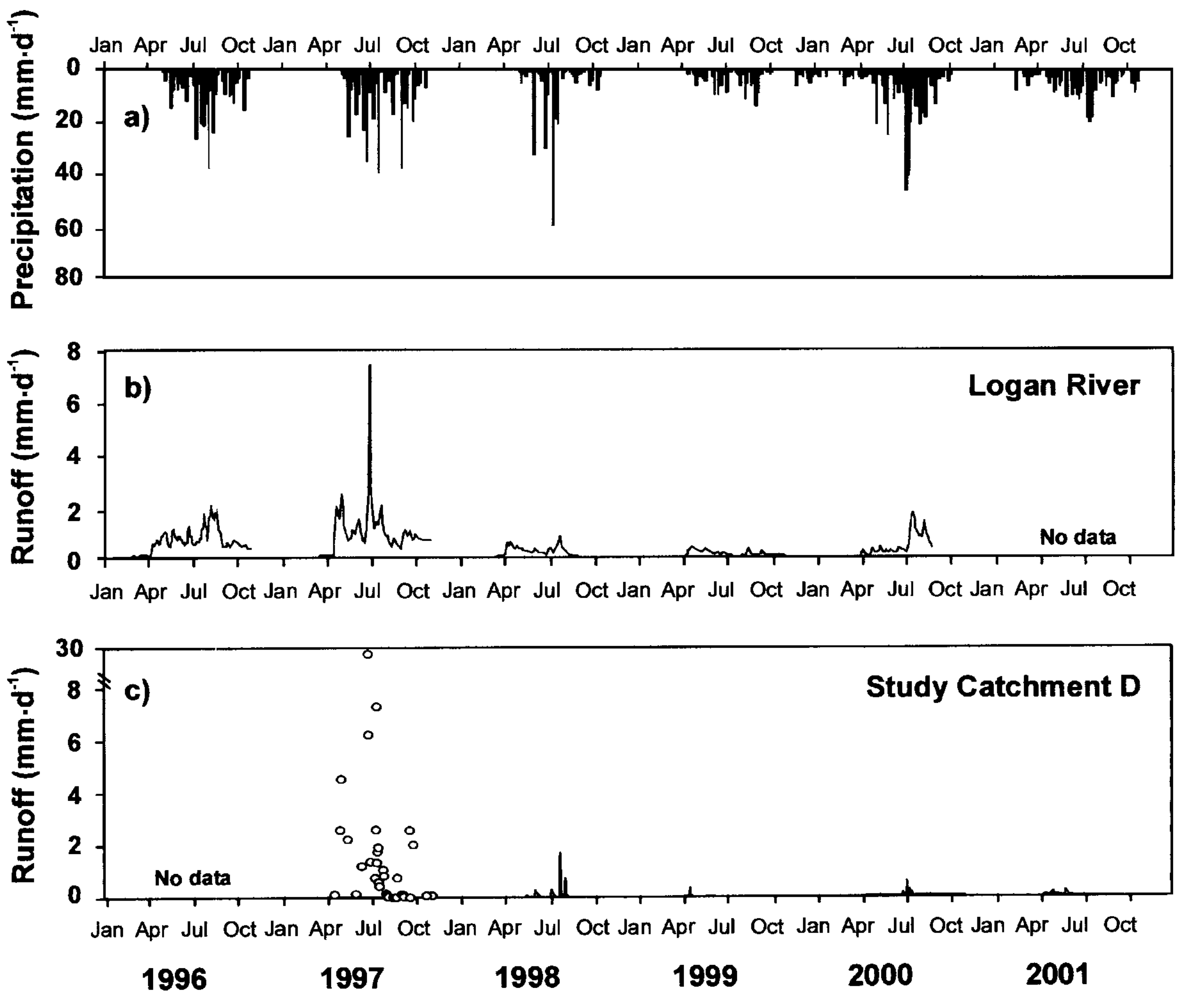

Figure 5. Daily (1996–2001) precipitation depths P and runoff from the Logan River and the study catchment

Subcatchments runoff comparison: potential influence of harvesting and stratigraphy

The importance of subtle variations in slope (hydraulic gradients), stratigraphy (depth to the confining layer), and soil frost of the ephemeral draws and valley-bottom wetlands on spatial variability in runoff among the subcatchments is illustrated during the wetting-up period of 2000 (Figures 7 and 8). Among the subcatchments, subcatchment B had a steep hydraulic gradient and a shallow depth to the confining layer that resulted in a large hydraulic gradient, a small water storage capacity, and a continuous discharge of water leading to significant subsurface and surface runoff (Table I, Figure 7). In contrast, subcatchment A had a steep slope but a greater depth to the confining layer, and no surface runoff was observed after 1997 (Table I). In subcatchment C, the depth to the confining layer varied, but the confining layer approached the ground surface near the outflow, where SSAs often formed with intermittent surface runoff responses (Figure 7). Similar to Boreal Shield catchments, hydraulic conductivity and depth of substrate over impermeable layers greatly influence the spatial and temporal patterns of stream discharge from headwater catchments (Hinton *et al.*, 1993).

Small runoff events occurred following snowmelt. Probing of the soil indicated that the organic-rich sediments in the lowlands of both subcatchments B and C were frozen, preventing infiltration and, thus,

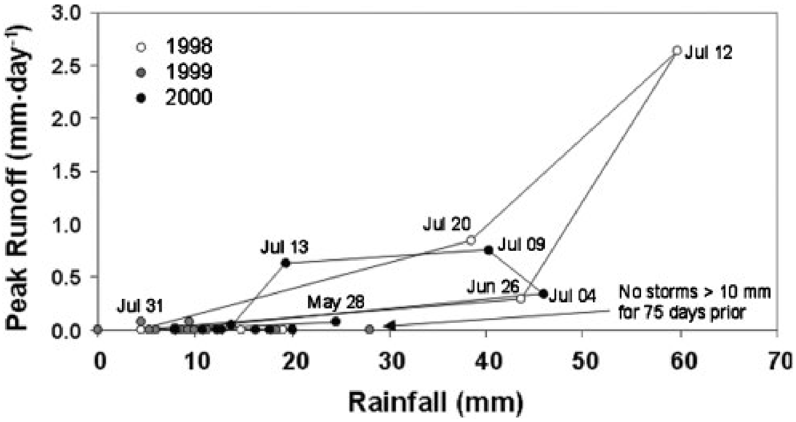

Figure 6. Hysteresis plots of precipitation event versus peak runoff from the catchment for selected storms in 1998, 1999, and 2000

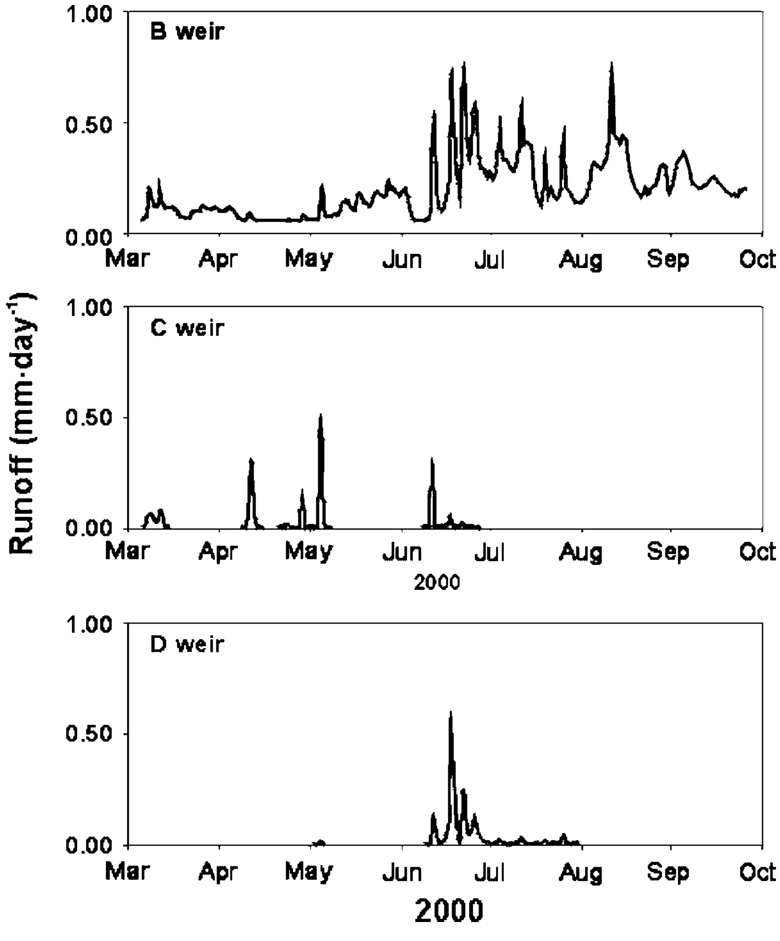

Figure 7. Daily surface runoff for the 2000 field season from the B subcatchment weir (top), C subcatchment weir (middle), and D catchment weir (bottom). No surface runoff was observed in subcatchment A

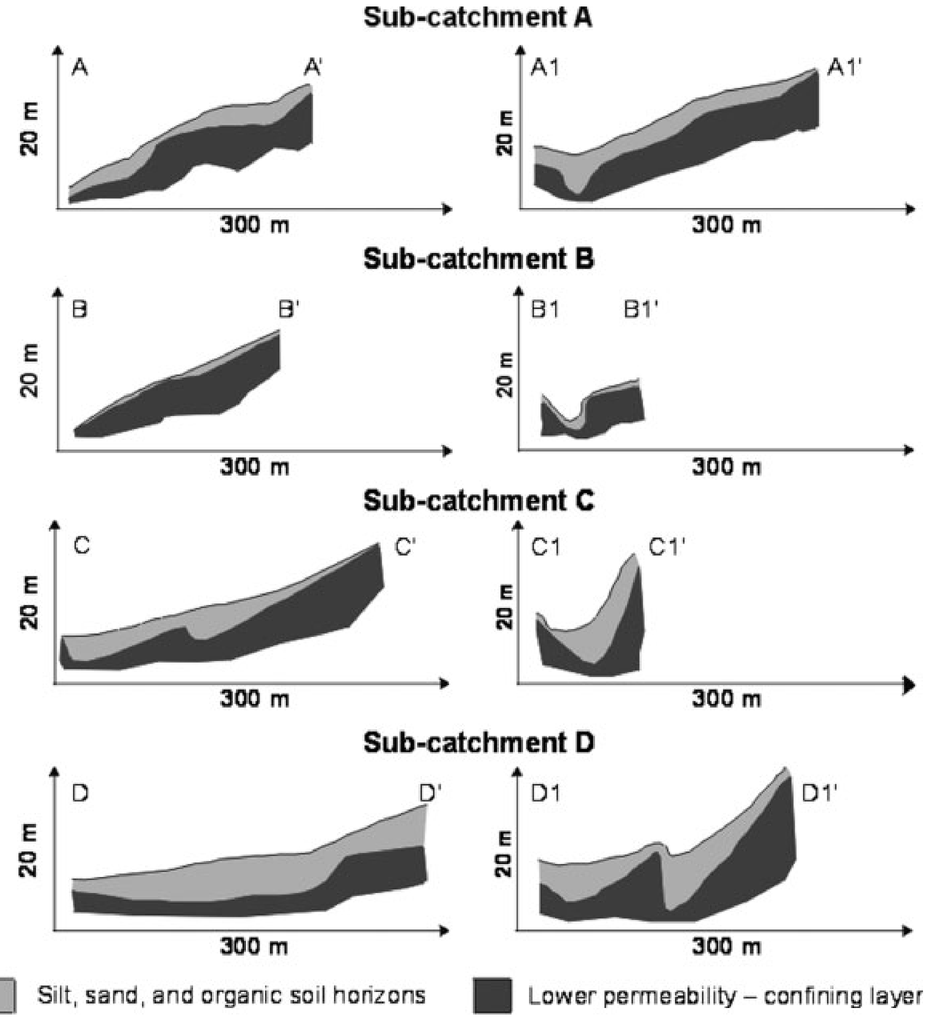

Figure 8. Relative elevation of the ground surface and the hydrologic confining layer along transects in the experimental catchment. Plots (a), (b), (c), and (d) correspond to transects in subcatchments A, B, C, and catchment D respectively (see Figure 2 for transect locations)

generating these runoff events (Kalef, 2002). However, runoff ceased shortly after soils thawed in subcatchment C. Subcatchment D had a moderate hydraulic gradient and the greatest depth to the confining layer. Here, no runoff events occurred following snowmelt. Probing of the soil indicated that the soils were not continuously frozen above the D-weir, and that snowmelt was lost to subsurface drainage. Several rainfall events in early summer 2000 were required to reduce the soil water storage capacity and to generate runoff at the C- and D-weirs (Figure 7).

Identification of the impacts of timber harvesting on the runoff regime of the catchment was confounded by large variations in precipitation following harvest, and spatial variations in soil storage capacity and groundwater interaction as influenced by subtle changes in the slope and the depth to the confining layer (Figure 7). Unfortunately, comparative measures of runoff were not made during peak runoff in 1997, which was the first season following harvest. In the years following, runoff was largest in subcatchment B, where approximately 88% of the area was harvested. However, the physiography of subcatchment B is conducive to runoff generation relative to subcatchments A and C (Figure 7; see Hinton *et al*. (1993)). The source areas of runoff were not restricted to the harvested areas. For example, a focused groundwater discharge site with permanent standing water occurred on the north hillslope adjacent to the valley-bottom wetland in the forested portion of the catchment (see Figures 2 and 13). Although no surface flow was observed, subsurface runoff from subcatchment A (34% harvested) was similar to combined subsurface and surface runoff from subcatchment C (73% harvested) during the second growing season following harvest (1998–2001, Table I). Furthermore, R and RCs were similarly low and decreasing in all subcatchments, regardless of area harvested (1998 to 2001), despite P in 2000 being 1·5 times greater than precipitation in 1998 and 1999 (Table I). Similar non-linear rainfall-runoff responses were observed in both harvested upper tributaries (subcatchments B and C) to that observed for the entire catchment, as illustrated in Figure 6 (Devito and Fraser, 2004).

The overall low runoff (Table I) and weak relationships between the area harvested and runoff from the subcatchment 2 years after harvest indicate that small precipitation depths in this region relative to soil moisture storage capacity may mask the influence of harvesting on catchment runoff. Logging occurred during the winter, resulting in minimal compaction of soils, with the exception of roadways (unpublished data). The rapid regeneration of aspen observed in the study catchment can potentially maintain high rates of evapotranspiration and high soil water storage potential, resulting in the low runoff observed (Westbrook and Devito, 2002). Although no measurements of AET in the harvested areas were made, potential rates measured using Penman methods in 2001 exceed precipitation (Devito and Fraser, 2004). It is uncertain whether tree harvesting effects on the Boreal Plain would be observed if aspen harvest occurred with significant soil compaction, or if harvest occurred in conifer stands, where the regeneration rates are lower than for clonal aspen.

Soil water storage

Independent measures of changes in soil water content and ground water-table levels further illustrate the influence of the soil water storage capacity on water fluxes. Although differences in the integrated soil water content in the top 1 m of the soil at the beginning and end of each water year were small (i.e. −3 mm in 1998, +22 mm in 1999, and −31 mm in 2000), the daily soil water contents ranged from 118 mm (29 March 1999) to 321 mm (10 July 2000) (Figure 9). The minimum soil water content (118 mm) followed two extremely dry years, and the maximum soil water content (321 mm) occurred when the soils were near fully recharged. The 200 mm range in soil water indicates that a large portion of the annual P can be stored in the soil, as observed in other aspen forest stands on the Boreal Plain (Blanken *et al*., 2001). These results emphasize the need for accurate characterization of soil properties and soil water content to predict differences in runoff from catchments.

Soil water content in the top 1 m of the soil profile of the hillslope at MetA (Figure 2) showed a consistent pattern of (1) a minimum in soil water content prior to spring snowmelt, (2) a recharge during spring snowmelt and rainfall, (3) a rapid depletion in early summer, (4) a subsequent recharge during summers with successive storms (i.e. 2000 and 2001), and (5) a rapid depletion following the summer storms through to early autumn (Figure 9). The snowmelt and rainfall that occurred in April and early May of each year accounted for the increase in soil water content in the spring. Infiltration experiments on frozen soils in the uplands of the catchment indicated that the frozen soils were porous and had a soil water storage capacity that was greater than the snow/water equivalent depth in most years (Kalef, 2002). Following the spring recharge, the reduction in soil water contents by evapotranspiration allowed for subsequent storage of the water from early summer storms and, thus, buffered the runoff responses to these storms in 2000 and 2001. Runoff responses were also buffered by the storage of water in the overstory and understory canopy and the litter–fibric–hemic (LFH) component of the soil, which on average was 8 cm deep. For example, at the MetA hillslope the LFH component stored about 15 mm of water, a water depth that is larger than many of the rainfall events occurring within each year (Figure 5). Recharge of the soils was often observed at the sand–clay interface (27 cm) and clayey-sand lens (35 cm), but less often observed in the sand below the clayey-sand lens (60 cm). Recharge to a depth of 60 cm was observed each spring, but significant recharge to 60 cm during the summers was observed only in 2000 (Figure 9).

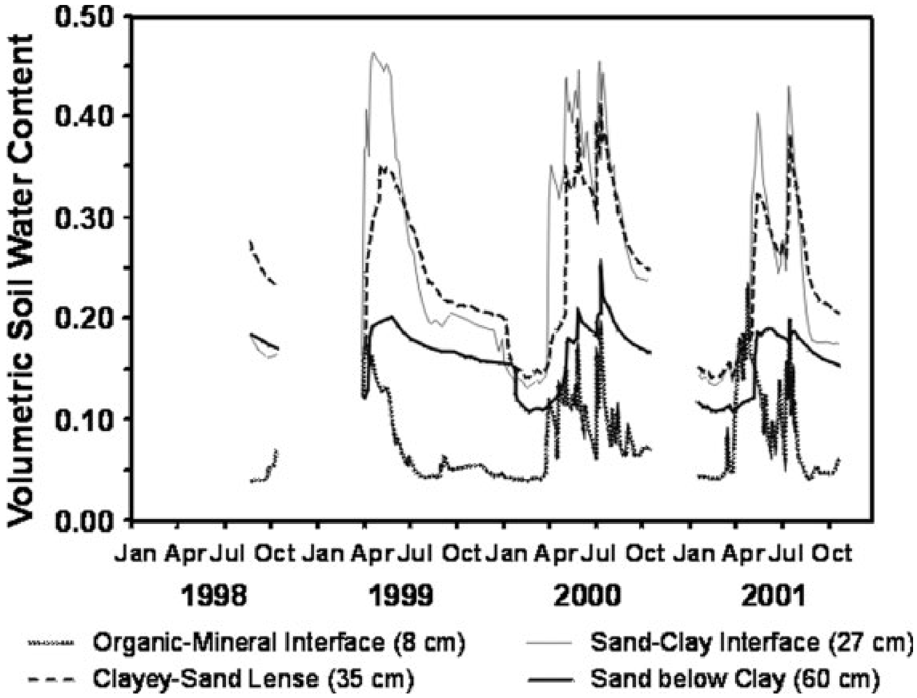

Figure 9. Volumetric water content at different depths in the soil profile for the aspen-forested hillslope, near MetB (see Figure 2): (a) organic–Ae interface (8 cm); (b) sand–clay interface (27 cm); (c) clay lens in sand (35 cm); (d) sand below clay lens (60 cm)

Rapid water table rises in upland wells suggest large, vertical, saturated hydraulic conductivities and, thus, recharge to depth (Figure 10a). Plots of rainfall versus water table responses show a linear response to rainfall events larger than 8 to 10 mm day^{-1} in the hillslopes (Figure 11a and b) and larger than 2 to 4 mm day^{-1} in the ephemeral draws and valley-bottom wetland (Figure 11c and d). The inverses of these slopes represent soil water holding capacities of 6 to 8% for the hillslopes and 10 to 11% for the wetlands. Assuming that the soil water holding capacities range from 6 to 11%, water table rises ranging from 40 to 80 cm (Figure 10 (2000)) represent 24 to 88 mm of groundwater storage. For April to July, cumulative water table rises ranging from 100 to 150 cm represent 60 to 165 mm of groundwater storage. Thus, a large proportion of rainfall entering the soils can be accounted for with a rise in the water table situated >1 m below the ground surface.

In the hillslopes, the water table was between 0·5 and 1·5 m below the ground surface in 1997, reflecting the potential for significant contributions from hillslopes to runoff in this year (Figure 10a). Although the water table was between 0·5 and 1·5 m below the ground surface early in 1998, it dropped to between 2·5 and 3·5 m below the ground surface by the end of the summer in 1998. In the remaining years, the water table showed a pattern of (1) lowest water table prior to snowmelt, (2) an increase in water table during spring snowmelt and rainstorms, and (3) a decrease in water table during early summer through to early autumn, with an intermittent increase in July when large rainfall events occurred. Some hillslope positions had water tables that rose to 0·5 m below the ground surface for a short period during the summer of 2000. The periodic rise in the water table is significant, as the saturated hydraulic conductivity, and thus the transmissivity of the soil, was larger near the ground surface. However, in general, the water table depth below the ground surface suggested small contributions from hillslopes to *R* during the study period (Figure 10a).

Compared with the uplands, the water tables in the ephemeral draws and the valley-bottom wetland had smaller seasonal fluctuations (Figure 10b). With the exception of the ephemeral draw in subcatchment A (W201), the water tables were at or above the ground surface at the initiation of spring melt and remained

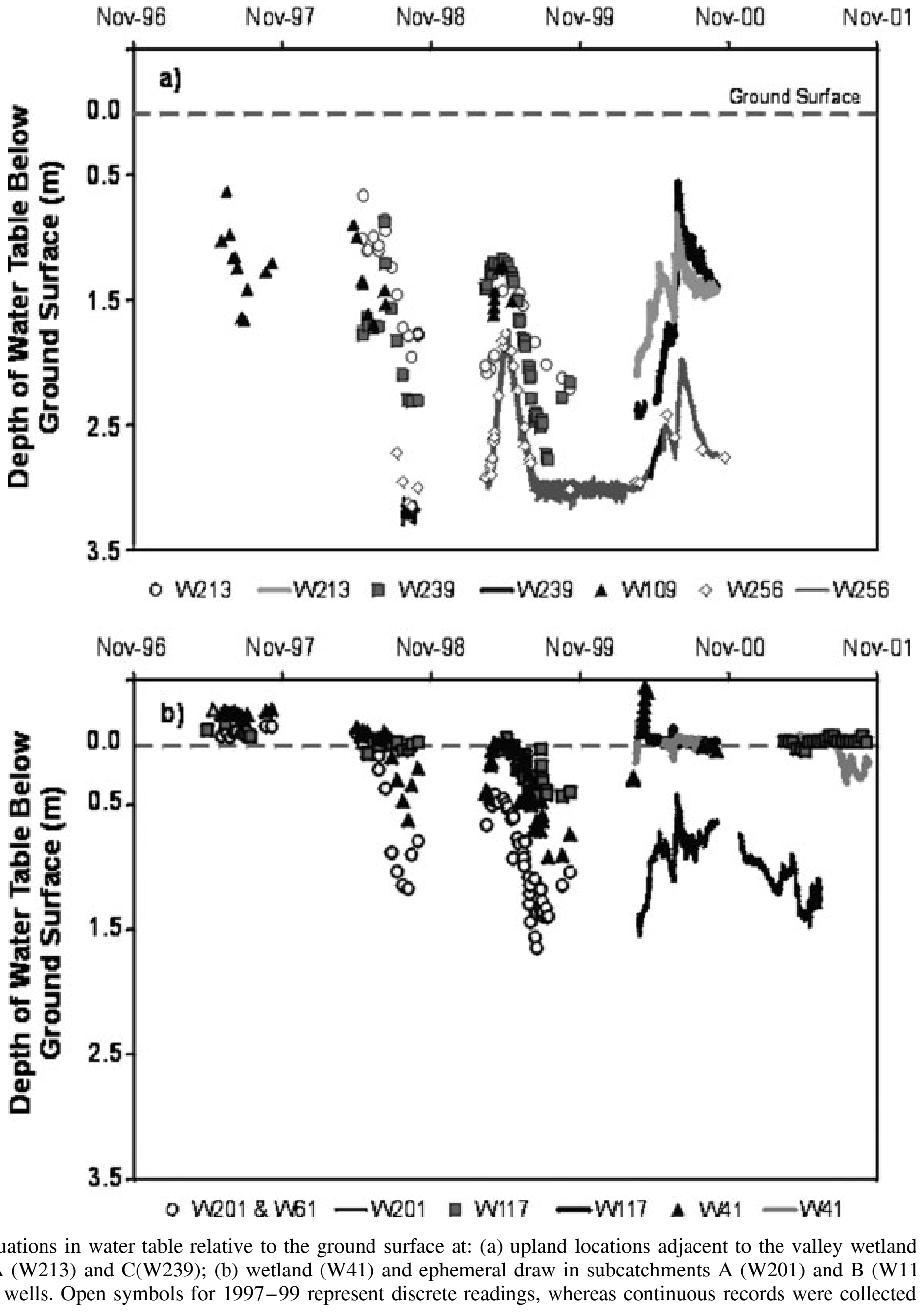

Figure 10. Fluctuations in water table relative to the ground surface at: (a) upland locations adjacent to the valley wetland (W109, W256), subcatchments A (W213) and C(W239); (b) wetland (W41) and ephemeral draw in subcatchments A (W201) and B (W117). See Figure 2 for locations of wells. Open symbols for 1997–99 represent discrete readings, whereas continuous records were collected 2000 and 2001

within 1 m of the ground surface throughout the rest of the growing season during the entire 5 year study. This minimal fluctuation suggested a larger potential to generate runoff here than in the hillslopes. There was standing water in all ephemeral draws and the valley-bottom wetland during the period in 1997 that coincided with high RCs from the catchment. The water table in the ephemeral draws and valley-bottom wetland dropped consistently in the late summer and early autumn with each successive dry year (1998 and 1999). The water table in the ephemeral draws of subcatchments B (W117) and C (Well101, Figure 12) and

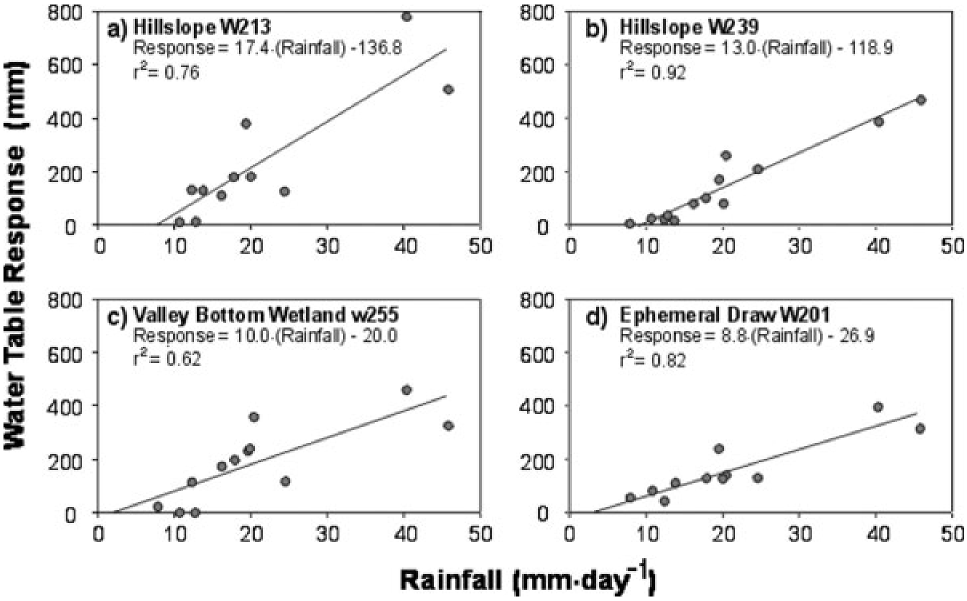

Figure 11. Plots of precipitation event to changes in the water table at different points in the study catchment: (a) hillslope (W213); (b) hillslope (W239); (c) edge of valley-bottom wetland (W255); (d) ephemeral draw (W201)

the valley-bottom wetland of D (W41) responded to summer recharge in 2000 and remained near the surface until late 2001 (Figure 10b). In contrast, the water table in the ephemeral draw of subcatchment A (W201) did not reach the surface during the summer 2000, and dropped to more than 1 m below the surface of the ground in 2001.

Deep water table responses are much greater than would be expected from the saturated hydraulic conductivities measured in the top 2 m of the soil throughout the catchment. Deep vertical recharge and lateral groundwater flow is restricted by low saturated hydraulic conductivities in the confining layer at the hillslope sites, but can account for approximately 15 to 50 cm drops in the water table on an annual basis. The remainder of the groundwater table recession must be accounted for by evapotranspiration or runoff along preferential flowpaths in the soil. The rapid water table rises, slow recessions, and negligible runoff response to rainfall indicate that subsurface runoff is restricted by the low transmissivity of the hillslope soils. Water table recessions were greater during the summer than the fall, indicating large draw down by evapotranspiration (Figure 10a). Excavation of soil pits throughout the study catchment indicate rooting depths of about 0·5 m, but coarse and fine roots were observed frequently to 2 m and as deep as 5 m, as reported in other aspen stands (Blanken *et al*., 2001). It appears that, in many years, the dual porosity of the glacial tills in the study catchment allows for water storage below the hydrologic active area, but stored water remains close enough to the ground surface to be affected by evapotranspiration. Soils with a high rate of vertical flow (recharge) and a low rate of lateral flow (runoff), resulting in storage of water and rapid rises in the water table, have been documented in prairie soils with similar surficial geology (Hendry, 1982; Winter and Rosenberry, 1995).

SSAs and their contribution to runoff

The formation of SSAs in the ephemeral draws and the valley-bottom wetland in response to rainfall events reflects: (1) a shallow depth to a confining layer; (2) a small water holding capacity in the organic-rich soils or sediments; (3) soil frost reducing infiltration; and/or (4) a contribution from groundwater flow systems.

The contributing area of groundwater required to maintain saturated areas of the size observed in wetlands and ephemeral draws would be small: <10% of the total catchment area. Identifying the source of the groundwater is problematic in the lowland areas due to the complexity of the groundwater–surface-water interactions, with concurrent gaining and losing sections and flow reversals. For example, in the north hillslope–ephemeral draw–south hillslope transect in subcatchment C, large hydraulic gradients were maintained between the north hillslope (W122) and the ephemeral draw (W121), indicating low transmissivity and small groundwater flow along this hillslope (Figure 12). In contrast, water table depths fluctuated significantly in the south hillslope (W239) and remained below the water table depths in the adjacent (W121) and downstream (W101) sections of the ephemeral draw during most of the study, with the exception of reversals in gradient during July recharge (Figure 12). Although the amount of groundwater recharge from the ephemeral draw to the south hillslope was probably small due to low soil transmissivities, its occurrence illustrates the complex hydrologic connectivity between hillslopes and lowland areas. Similar hydrologic interactions were observed between the hillslope and confluence of subcatchments A and B (data not shown) and between hillslopes and the valley-bottom wetland (Kalef, 2002). Other studies have observed similar hydraulic gradients and flow reversals in forested uplands associated with wetlands (Taylor and Pierson, 1985; Devito *et al.*, 1997) and ponds in glaciated areas (Hayashi *et al.*, 1998; Ferone and Devito, 2004).

The water table gradients indicate that the hillslopes are largely hydrologically disconnected from the valley wetlands and ephemeral draws. Consistent hydraulic gradients down the valley areas (Figure 12) and slightly higher hydraulic conductivities of valley soils/sediments indicates that up-catchment valley areas are most likely the groundwater contributing areas, and that the ephemeral draws and valley-bottom wetland are the major sources of runoff.

The SSAs in the catchment that existed during the week of peak daily runoff in each year were positively correlated with the peak and the total annual runoff (Figure 13). More than 8% of the catchment (SSA = 42 970 m^2) was saturated following the large rainfall and runoff events in 1997 (Figure 13A). The distribution of the SSA followed the topographic lows and road rights of way. The SSAs were hydrologically connected, and surface runoff flowed along SSAs from the ephemeral draws to the valley-bottom wetland and

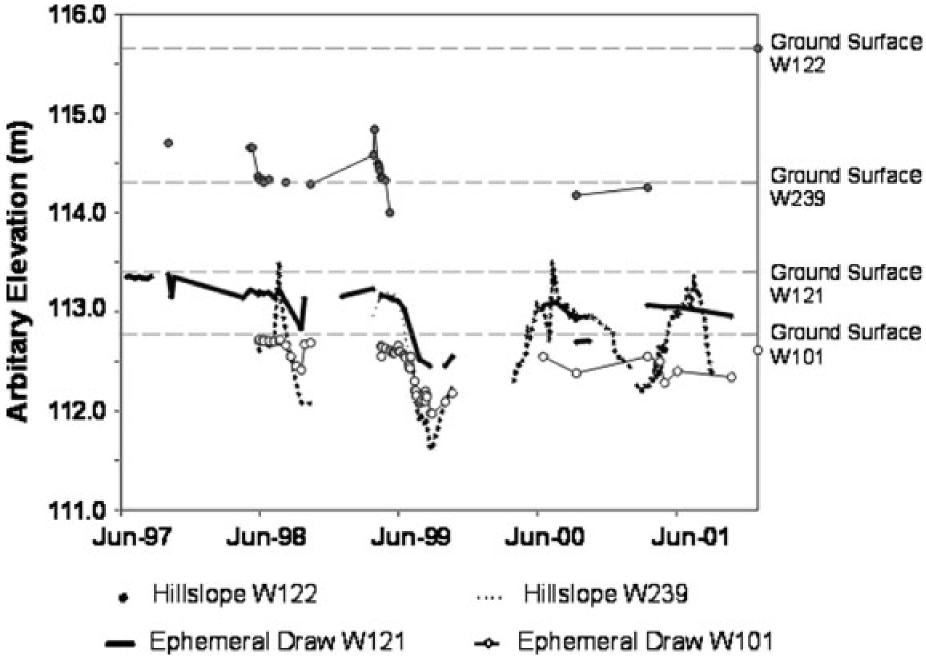

Figure 12. Water level elevations in wells on the north hillslope (W122), south hillslope (W239) and upstream (W121) and downstream (W101) locations of the ephemeral draw above subcatchment C weir. Dashed lines reference ground elevations at the base of each well

to Moose Lake in each subcatchment during this time. About 4% of the catchment (SSA = 16 169 m^2) was saturated during peak runoff in 1998 (Figure 13B). The SSAs became hydrologically disconnected in the upper reaches of the catchment and in subcatchment A. The reduction in SSAs that were hydrologically connected to Moose Lake was related to the reduction in *R*. Following the drought in 1998 and 1999, there was a dramatic reduction in the SSAs, with SSAs maintained only in areas of perennial discharge in subcatchments B, C, and D (SSA = 343 m^2). There was negligible peak runoff during this time (Figure 13C). The above-average rainfall in 2000 recharged the ephemeral draws and valley-bottom wetland (SSA = 10 264 m^2). Although the SSA increased, the SSA remained below that of 1998, was discontinuous, and *R* was small (Figure 13D).

The SSAs were seasonally dynamic (Figure 14). For most years following 1997, the maximum SSA that was hydrologically connected to the subcatchment and catchment outflows occurred in April and May, when *R* was negligible (Figure 5). High water table or standing water was observed above the weirs in each year during and after snowmelt (Figure 10b). Groundwater flow reversals, with water tables higher in the lowlands

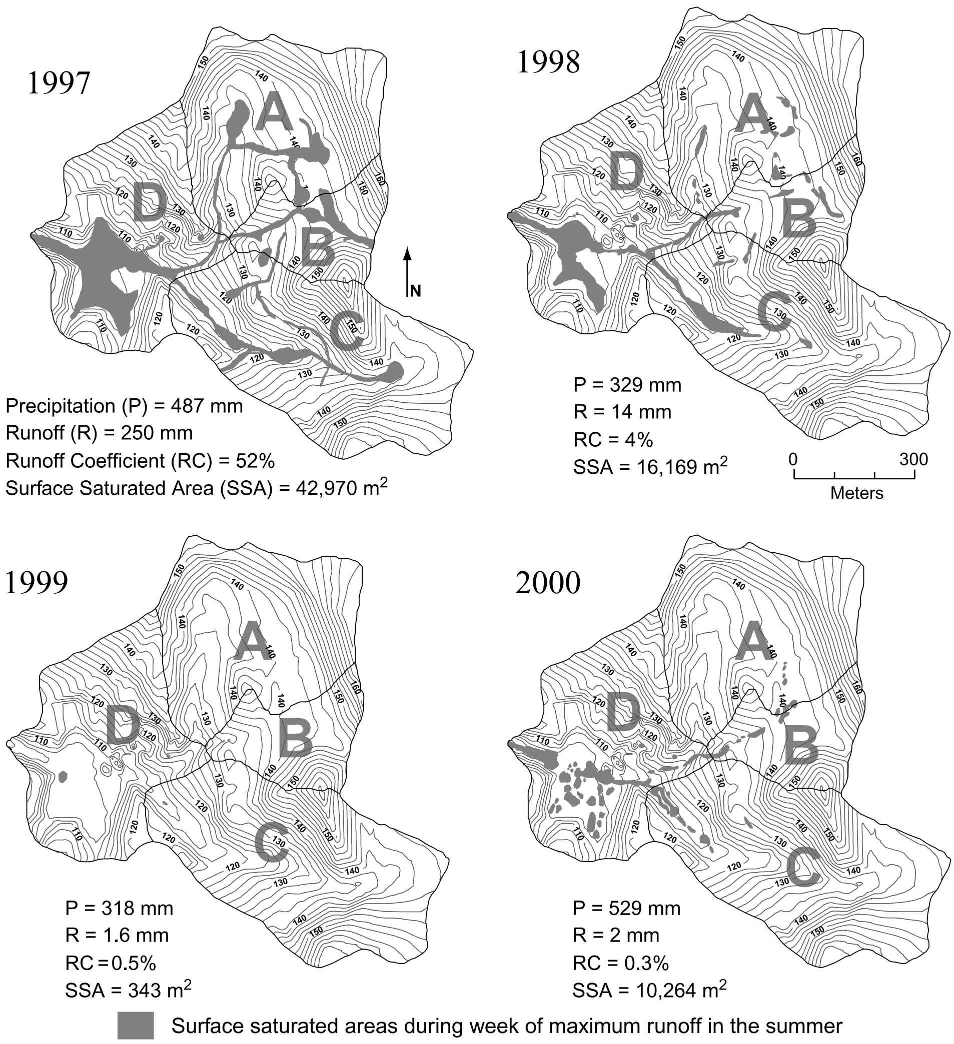

Figure 13. Maps of SSA for the week of maximum daily runoff in July of (a) 1997, (b) 1998), (c) 1999, and (d) 2000. Annual precipitation *P*, runoff *R*, and runoff coefficient RC for each year are provided

than the adjacent uplands, were observed in several areas, indicating the presence of a perched water table in the ephemeral draws and valley-bottom wetland. Investigation of the depth to ice and the infiltration rates into the soils suggest that organic-rich soils in the ephemeral draws and in the valley-bottom wetland were frozen and impervious prior to spring melt (Kalef, 2002). Although the potential for peak runoff existed, accumulated snow depths were small and the probability of rainfall in April and early May was very low (Table II); thus, little runoff was generated.

Once soils thawed in May, there was a rapid decrease in the SSAs. With the exception of 1999, successive storms in late June and July increased the extent and connectivity of SSAs. Evapotranspiration, coupled with infrequent rainfall, resulted in rapid reduction and disconnection of saturated areas in late summer and early autumn. By the end of the summer, the only remaining SSAs were located at the base of the weir at subcatchment B and C and at a focused recharge area on the north hillslope above the valley wetland in catchment D.

Analyses of the ratio of R to the SSAs that were hydrologically connected to Moose Lake emphasized the importance of temporal and spatial variations in SSAs as potential source areas of R. Instantaneous measures of the extent and connectivity of SSAs during storms were not possible, but the rapid rise and recession of water table depths in the lowlands suggested that there was a rapid expansion and contraction of SSAs. In July 1997, the SSAs probably approached an upper limit for the catchment (Figure 13A). Thus, RCs calculated by

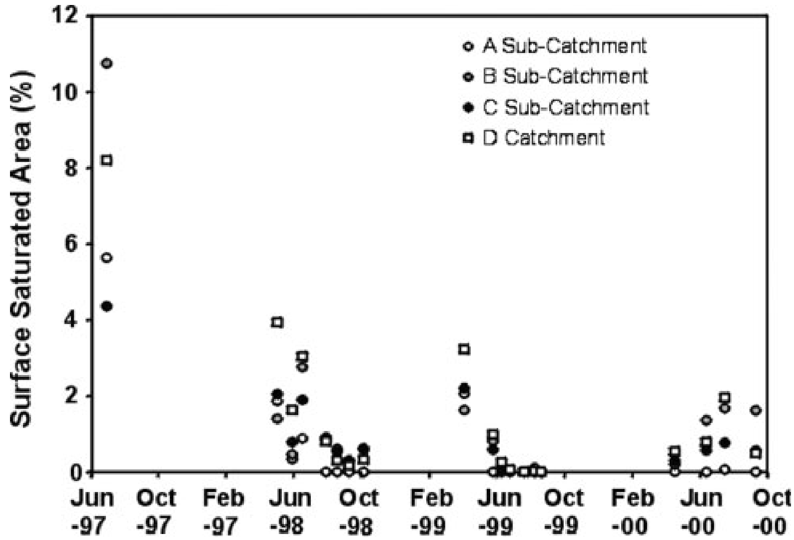

Figure 14. Percentage of SSAs within the A, B, and C subcatchments and the entire catchment D for selected dates (1997–2000)

Table II. Probability of a precipitation event exceeding a specified threshold and the average daily PET for the months of April to October for 1995–2000

Month	$P \geq 10$ mm	$P \geq 30$ mm	PET (mm day^{-1})
April	0·010	0·000	2·1
May	0·046	0·005	3·6
June	0·094	0·020	4·0
July	0·110	0·028	4·0
August	0·063	0·009	3·0
September	0·058	0·010	1·7
October	0·009	0·000	0·8

Table III. A comparison of RCs measured at the B, C subcatchments and D catchment weirs based on total area (CA) versus the SSA in the catchment that is hydrologically connected via surface flowpaths to Moose Lake. Selected precipitation events during the 1997, 1998, 1999 field seasons and all precipitation events during the 2000 field season are provided

Year	Date	*P* (mm)	B subcatchment		C subcatchment		D catchment	
			RC/CA	RC/SSA	RC/CA	RC/SSA	RC/CA	RC/SSA
1997	8 July	35·0	38·7	660·1	84·5	1050·7	53·4	711·5
1998	1 Jun	30·2	—	—	—	—	4·3	58·2
1998	27–28 Jun	64·6	5·3	90·3	—	—	2·0	27·5
1998	20 Jul	27·0	22·0	374·4	—	—	5·4	73·1
1999	4 Jul	23·6	0·4	6·4	0·0	0·0	0·0	0·0
1999	1 Sep	36·0	0·0	0·0	0·0	0·0	0·0	0·0
2000	4 May	20·0	0·2	2·4	2·1	54·1	0·0	0·0
2000	28 May	24·5	0·7	7·6	2·1	55·4	0·5	7·1
2000	4 Jul	46·0	1·2	12·7	0·6	16·3	0·4	4·9
2000	9 Jul	40·3	1·5	16·8	0·1	3·8	2·3	31·5
2000	13 Jul	19·4	3·3	36·1	0·1	2·4	3·3	44·4
2000	1 Aug	20·0	1·6	17·0	0·0	0·0	0·1	1·9
2000	15 Aug	17·8	1·4	15·1	0·0	0·0	0·2	2·8

the ratio of R to this upper limit for SSA ($R:\text{SSA}$) were compared with the RCs calculated by the ratio of R to catchment area ($R:\text{CA}$) (Table III). Following 1997, the $R:\text{SSA}$ ranged from 0 to 90%, with the exception of one summer storm in July 1998 ($R:\text{SSA} = 374\%$). The RCs based on $R:\text{SSA}$ support the observed soil water content and water table depth patterns, and suggest that R is generated largely from SSAs in ephemeral draws and valley-bottom wetlands.

Hydrologic disconnection of uplands and adjacent wetland or ephemeral draw regions in headwater catchments has been observed in a range of climatic and geologic regions (Devito *et al*., 1996; Buttle *et al*., 2000). This research further supports recent reviews suggesting that riparian areas or ephemeral draws and adjacent forest uplands be defined as discrete hydrologic response units in modelling headwater catchment soil moisture distribution and rainfall-runoff responses (Grayson and Western, 2001; McDonnell, 2003).

CONCLUSIONS

In aspen-dominated catchments located on the subhumid Boreal Plain, P–evapotranspiration deficits are equal to or greater than soil water storage capacity in most years (also see Blanken *et al*. (2001)), and spatial and temporal variations in soil water storage exert a large influence on storm and annual runoff regimes. Thus, storm and annual runoff is controlled more by the distribution of P in time than seasonal or annual totals in P. The largest rainfall events occur in July. However, this occurs when evapotranspiration is highest, and the resulting soil water storage capacity buffers runoff from storm events. Successive rain events >30 mm within 1 week are required to exceed soil water storage capacity and produce runoff, but the probability of this occurring is low. As observed in recent studies on the Boreal Plain (Gibson *et al*., 2002; Wolniewicz, 2002), wetlands and ephemeral draws with organic-rich soils are important runoff-generating areas. These studies also show that decoupling of uplands from wetlands occurs in most years, and each system may be considered as a discrete hydrologic response unit (Grayson and Western, 2001; McDonnell, 2003). Disconnection of ephemeral draw or wetland source areas leads to low total runoff from headwater catchments, as well as from the Boreal Plain landscape (RC < 25%, generally less than 100 mm). During wet years, the soil water storage capacity can be exceeded, resulting in the coupling of uplands to wetlands and catchment outlets. Such events occur infrequently (every 10–20 years) and may be difficult to predict, but runoff may be large (>200 mm, RC > 50%).

Regional comparisons indicate that runoff from Boreal Plain catchments is intermediate to runoff generation in: (1) subhumid, subarctic catchments in the Yukon, where runoff generation is restricted to frozen organic soil areas, and does not occur from aspen-dominated areas free of permafrost where melt water and rain go into soil storage or are lost to evapotranspiration (Carey and Woo, 1998, 1999); (2) humid eastern Boreal catchments, where coupling and decoupling of hillslopes with valley wetlands influences seasonal variation in catchment runoff (Devito *et al*., 1996; Buttle *et al*., 2000). Runoff due to snowmelt in the Boreal Plain is small compared with the Boreal Shield (Devito *et al*., 1996) and the Boreal Cordillera (Carey and Woo, 1998). The probability of rainfall in late summer and fall is very low, resulting in low soil water storage prior to soil freezing in Boreal Plain catchments. These conditions greatly reduce spring snowmelt responses in glaciated regions (Granger *et al*., 1984; Woo *et al*., 2000). Furthermore, although frozen valley wetland soils and snowmelt primed the study catchment for runoff, low total snow accumulation, discontinuous frost layers in wetland areas coupled with high soil storage in hillslopes, and a very low probability of spring rainfall resulted in little runoff during the snowmelt period.

The influence of tree harvesting on runoff at the study catchment was inconclusive because of large interannual variation in rainfall and spatial variability in lithology and soil storage. Nevertheless, runoff from the study catchment (50% of the forest harvested) and from two tributaries (>75% harvested) was low, similar to or less than regional runoff 2 to 4 years following harvest. There is a low potential for tree harvesting to affect water yields at this site due to the large P−evapotranspiration deficit relative to soil water storage, low frequency of large storms (>30 mm day^{-1}), and low runoff coefficients. Low spring runoff on the Boreal Plain further reduces harvesting impacts compared with eastern Boreal catchments, where spring snowmelt can represent a large portion of annual runoff and may magnify impacts from disturbance such as forest removal (Buttle *et al*., 2001).

Most runoff appears to occur from a small proportion of the catchment; thus, the valley-bottom ephemeral draws and/or wetlands may be susceptible to harvesting impacts during most years. Taking care not to enter these areas and to avoid creating hydrologic connections via road networks is important to minimize impacts on water yields.

This study illustrates that site selection for harvesting studies and interpreting results from paired catchments in this climatic and physiographic region is problematic. The deficit between P and evapotranspiration is low and similar to the storage capacity of the soils. Furthermore, the variabilities in both soil water storage capacity (required to be exceeded for runoff) and interannual variation in P−evapotranspiration deficit are large. Thus, the soil water storage capacity at one catchment may be exceeded more frequently than at an adjacent catchment due to subtle differences in soil characteristics. Long-term monitoring of runoff responses that result from different annual and seasonal rainfall and evapotranspiration patterns, and/or spatially extensive soil characteristic measures, are required to determine catchment similarity prior to experimental manipulation. Long-term climate records indicate that this period could be as long as 10 years on the Boreal Plain, which is the length of time required for aspen regeneration to recover to pre-harvest transpiration capacity.

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